

The Arctic Ocean boundary layer: Interactions with the sea-ice surface and clouds

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Abstract

The last several decades has seen a large climate change in the Arctic with a warming twice or more as large as the global average. As a consequence there have been many changes in the Arctic environment, most pronounced in the thickness and extent of the perennial sea ice. The ice cover has decreased by about 10% per decade and the last few years show an accelerated decrease. The lowest end-of-summer melt season ice cover, usually observed in mid-September every year, occurred in 2007 and every year since 2007 has had a lower ice cover extent than any of the years before 2007, during the some 30 years of satellite observations. At the same time the amounts of thick so-called multi-year ice has decreased even faster. Numerical models used either to understand Arctic change, for example using reanalysis products, to forecast Arctic weather for those living in the Arctic or to project the future state of the Arctic climate, have significant problems dealing with processes in the Arctic, even under current conditions. This has several reasons; one major problem is the paucity of detailed process-level observations. As a consequence many processes unique to the sea-ice covered Arctic Ocean are poorly understood and therefore poorly parameterized in models. The lack of observations in general also contributes to difficulties in evaluating models for the Arctic Ocean. Among the special features that are poorly handled by models are the very stable conditions sometimes occurring in winter and the often occurring low-level mixed-phase clouds, with liquid water present at very low temperature. The transfer of heat through the ice and snow in winter is also poorly handled by models, as are the effects on surface albedo by melt ponds from melting snow on the ice surface in summer. All of these processes have pronounced effects on the surface energy balance and hence on the melting and freezing of the sea ice and the energy balance at the surface and Arctic boundary layer processes hence lies at the heart of this change.

1. Introduction

The Arctic is at one of the cold end of the climate heat engine; excess heat from solar radiation in equatorial regions is transported by the ocean and the atmosphere to the Arctic where it is lost to space by radiation. Hence an understanding of the energy balance of Arctic is central to understanding the global climate. The surface conditions in the Arctic Ocean are unique, with its perennial sea-ice cover. The surface energy exchange processes are of central importance to the annual cycle of sea-ice melt and freeze. Another feature of great importance is the large annual cycle due to the high latitude; in winter the sun is essentially absent for a long period and the sea ice increases in areal extent and thickness, while in summer the sun is over the horizon the entire day and the sea-ice melts as a consequence of a positive surface energy balance. Meanwhile, the temperature at the sea-ice/ocean interface remains at the freezing point for saline water, at about $-1.8\text{ }^{\circ}\text{C}$, and the ice thickness varies in response to the surface energy balance, while open ocean liberated from ice in summer can absorb and store significant amounts of heat into following season. The Arctic Ocean is also a very cloudy place; in winter the cloud fraction drops to 60-80% in response to the lower surface water-vapour saturation over ice and snow, while in summer it often reaches $> 90\%$. Much of the clouds are low-level mixed-phase stratocumulus in a shallow boundary layer that is neither continental nor marine in character; it is special for the Arctic Ocean.

The harsh environmental conditions especially in winter, the fact that the ocean is ice-covered and that the ice continually drifts make observations of the atmosphere difficult; stationary observatories are in principle impossible and even drifting observatories are challenging. As a consequence our understanding about these very special conditions is very limited. Most comes from either sporadic ice-breaker based expeditions, networks of surface observations from drifting buoys, from a few drifting so-called “ice-islands” and from satellite observations. Passive satellite observations also have significant problems distinguishing clouds from the snow-and-ice surface and retrieval of atmospheric parameters are limited due to uncertainties in surface emissivity. The only existing detailed observations covering a full annual cycle comes from the Surface Heat and Energy balance of the Arctic (SHEBA) expedition from September 1997 to September 1998 (Uttal et al. 2002) on a drifting icebreaker in the Beaufort and Chukchi Seas. Detailed summer expeditions are more numerous, for example the International Arctic Ocean Expedition in 1991 (IAOE-91, Leck et al. 1996), and the Arctic Ocean Experiments in 1996 and 2001 (AOE-96, Leck et al. 2001; AOE-2001, Leck et al. 2004; Tjernström et al. 2004) and the Arctic Summer Cloud Ocean Study (ASCOS, e.g. Sedlar et al. 2011a). Consequently, we know less about Arctic Ocean boundary-layer processes than in any other region on Earth and what we know is heavily biased to summer conditions.

2. The Arctic lower troposphere vertical structure

The boundary-layer is much influenced by the vertical structure of the lower troposphere that governs the interplay between the surface and the free troposphere. There is a common misconception about the Arctic, that it is a place dominated by very stable stratification. Tjernström and Graversen (2008) conducted an objective analysis of low troposphere inversions from the SHEBA soundings and also compared to the European Centre for Medium Range Weather Forecasts ERA-40 reanalysis. Figure 1 shows the height dependent probability of the vertical gradient of the potential temperature. While there are differences in details between summer and winter, the structures are surprisingly similar. Although strongly stable conditions near the surface are more common in winter than in summer, the statistically dominant feature is a well-mixed boundary layer capped by an inversion around a few hundred meters. All temperature profiles from SHEBA were analyzed for inversions using an

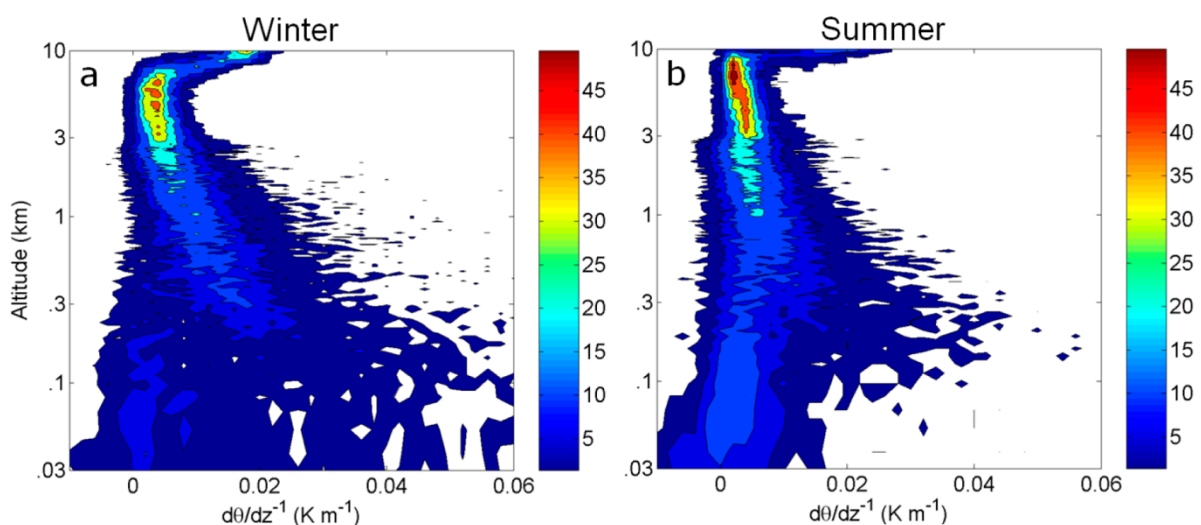


Figure 1. Probability (color) of the vertical potential temperature gradient (Km^{-1}) for (a) winter and (b) summer calculated from all SHEBA soundings.

objective algorithm (Tjernström and Graversen 2008). Defining “surface inversions” as those with a main inversion-base height < 25 m, all temperature profiles were then categorized as having either “surface” or “elevated” inversions; all profiles except one(!) had at least one inversion.

Figure 2 shows the normalized temperature structure. Here the height of the lowest layer is normalized by its depth: using either the height to the top of the inversion, for the surface-based inversions, or the height to the inversion base, for elevated inversions. In both cases the temperature is scaled by the total temperature difference across the lowest layer. It is clear that two distinct regimes exist, one with a surface-based inversion of varying strength, with a median around $\Delta T \sim 10^\circ\text{C}$, and one with an elevated inversion below which the temperature decreases linearly with height – a well-mixed boundary layer. While the surface layer is more often stable in autumn and winter, occurring about 61 and 53% of the time during the SHEBA year, spring and summer had almost no stable conditions, only 15 and 9% of the time. The latter seasons are thus totally dominated by well-mixed but shallow boundary layers and even in autumn and winter these conditions occurred 39 and 47% of the time, respectively. Seen over the whole year, well-mixed conditions thus dominated and even in the winter season, well-mixed conditions were frequent.

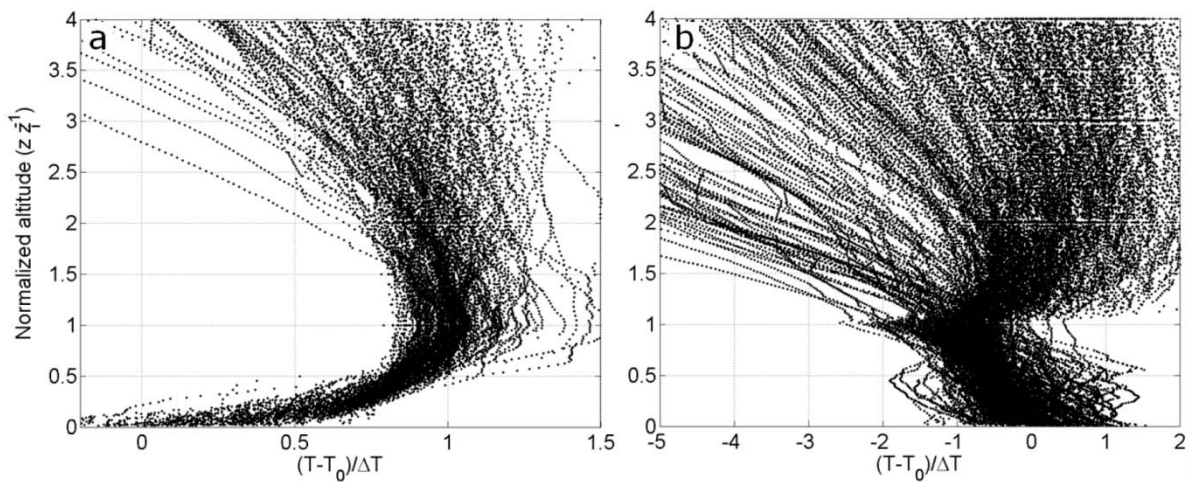


Figure 2: Scaled boundary-layer temperature structure from all SHEBA soundings for (a) surface inversions and (b) elevated inversions. The categorized are simply distinguished by a threshold of 15 m for the height to the base of the inversion

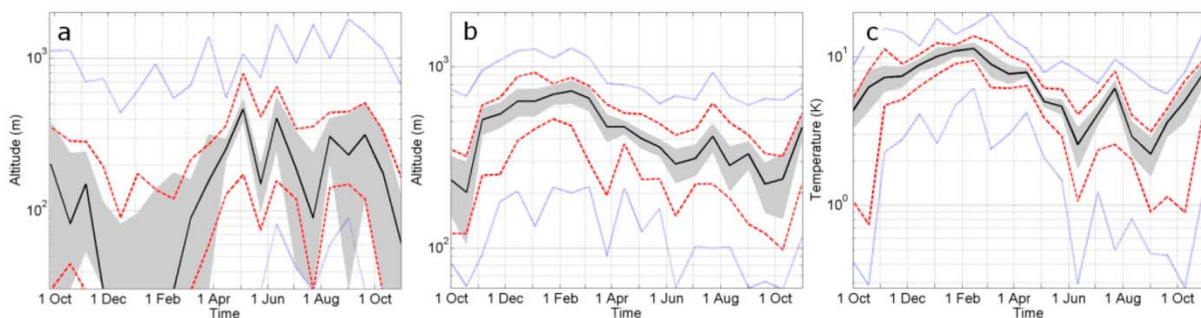


Figure 3. Annual cycle of inversion (a) base height and (b) depth (m), and temperature strength ($^\circ\text{C}$) evaluated over 20-day periods during the SHEBA year. The solid black line shows the median while the gray shaded area shows the double sided 95% significance. Red dashed lines show the 25- and 75-percentiles while blue dashed lines show the 5- and 95-percentiles.

Figure 3 shows the annual cycle of some inversion metrics; the height to the inversion base, the thickness of the inversion and the temperature jump across the inversion. While low inversion bases dominate in winter, the variability is also quite large and up to ~200 m high inversion bases occur frequently. The inversions are stronger and deeper but also more variable in winter, while summer conditions are less variable, with typical inversion base heights around a few hundred meters with and shallower and weaker boundary-layer capping inversions.

In summary, while very stable conditions do occur in autumn and winter, the most commonly occurring structure of the Arctic boundary layer is a shallow and near-neutral layer with a capping inversion. In spring and summer the latter totally dominates, while in autumn and winter conditions “flip-flop” between well-mixed near-neutral conditions and surface-based inversions. Not surprisingly, these two regimes correspond to cloudy and clear conditions, where clear in this context is defined from the radiative characteristics so that optically thin clouds, with a small impact on the surface net radiation, are considered “clear”. For much of the year, low-level clouds tend to warm rather than the surface (e.g. Intrieri et al. 2002).

3. Some basic features near surface features

Before proceeding with detailed discussions it is instructive to examine some basic meteorological parameters and how they appear in the different seasons. Figure 4 shows 60-day time series of near-surface temperature for winter, spring and summer from SHEBA. In winter (Figure 4a) temperatures are generally low, roughly between -20 and -40 °C, and changes rapidly between these two “end states” in a seemingly random fashion. In winter the sun is below the horizon and so only longwave radiation, turbulent fluxes and heat conduction in the snow and ice determines the surface temperature; of these the longwave radiation fluxes are by far the largest and also vary the most. The most important factor determining the net longwave radiation at the surface is the low-level clouds. Thus the cold end of the scale represents clear conditions and also features the strong surface inversions, see above, while the less cold state reflects the cloudy and well-mixed boundary-layer conditions. It is also interesting to note that in winter, the temperature difference between the ice surface and the atmosphere, across the snow pack, is much larger than between the ice surface and the underside of the ice, which must remain at freezing (~-1.8 °C) even though the thickness of the ice is typically much larger than the thickness of the snow.

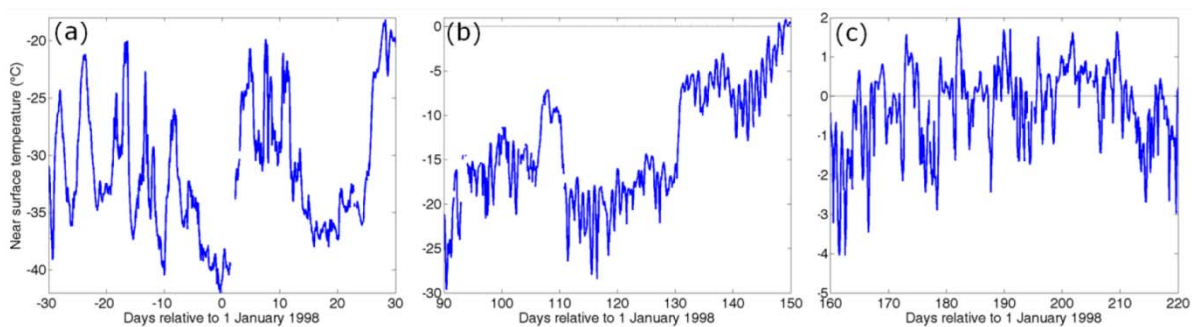


Figure 4. 60-day time series of near-surface air temperature (°C) for showing characteristic (a) winter, (b) spring and (c) summer conditions

In spring (Figure 4b) the temperature is also highly variable, roughly between -30 and 0 °C, but now the larger part of this span is determined by the gradual warming trend in the months of transition from winter to summer. The most prominent temporal variability arises from a distinct diurnal cycle. The sun is over the horizon part of the day and warms the surface during daytime, and while the surface is still below freezing the temperature responds much like a land surface. Finally, in summer (Figure 4c) the temperature fluctuations are much smaller than in any other season, mostly ± 1 °C around zero. The sun is now up the entire day, but the excess energy goes into a melting of the surface rather into changing its temperature, thus no clear diurnal cycle can be observed.

Thus three distinct regimes can be observed: 1) Winter, when surface temperature, and boundary-layer structure, is determined by the presence of low-level clouds; 2) Spring, when the sun is strong enough to enforce a diurnal cycle, as well as a gradual warming, but the temperature is below zero and is still free to vary in response to radiative forcing; 3) Summer, when there is an excess of surface forcing but all this energy goes into surface melt. The summer surface melt in itself also increases the surface forcing in a positive feedback, involving both the exposure of open ocean water and the formation of melt ponds in the ice surface. This decreases the surface albedo and leads to even more absorption of solar radiation; the famous surface albedo feedback (e.g. Perovich et al. 2008).

The distinction in the surface energy balance between winter and spring on the one hand, when the surface temperature is free to respond to the changes in the surface radiation balance, and summer when it is not, is clearly illustrated in Figure 5 showing the relationship between radiation and turbulence/conduction surface energy fluxes in “winter” and “non-winter”. This reflects the ability of the surface temperature to respond to changes in the net radiation; this ability is obviously impaired during the melt season. While the net radiation changes over roughly an equal range in both seasons, the other surface heat fluxes respond very differently. The near surface relative humidity (not shown) is constantly high. The relative humidity is almost always close to saturation; in winter with respect to ice. This also partly explains for the lower cloud fraction in winter compared to summer; the low clouds often consist of liquid droplets while the boundary layer moisture is determined by ice saturation.

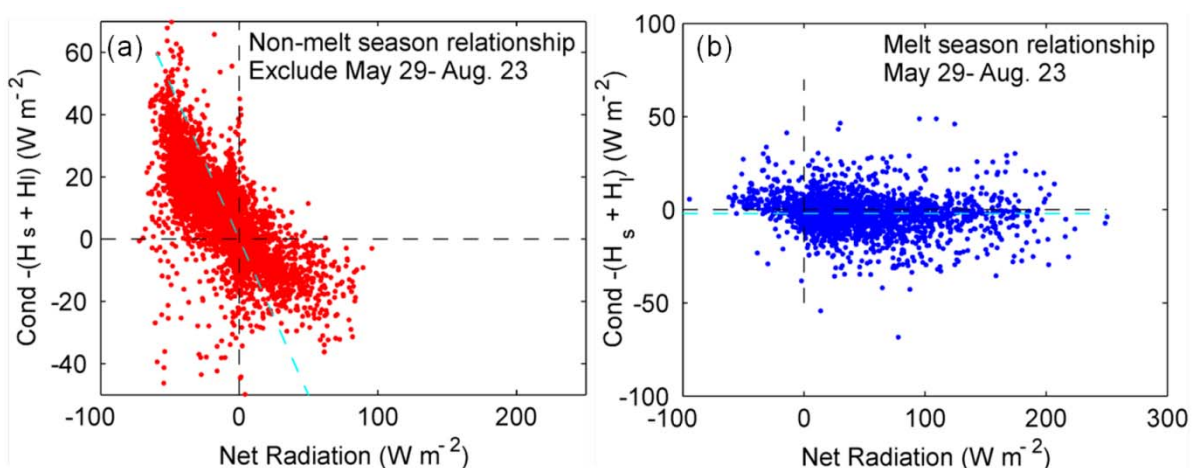


Figure 5. Scatter graphs showing the relationship between the sum of the turbulent heat fluxes and the heat conduction through the ice and the net radiation, both in W m^{-2} , at the surface for the (a) non-melt and the (b) melt seasons. Reproduced from Persson (2011).

4. The winter-time boundary-layer

Figure 6 shows the vertical structure of the lower troposphere winter sub-sampled such that cloudy and clear cases are averaged separately; two things are immediately clear. First, the cloudy cases are warmer, not just in the boundary layer but through the whole lower troposphere. This shows that for clouds to be prevalent, warmer temperatures are in general also present and suggests that these periods may be connected to an inflow of warmer air from lower latitudes and thus a synoptic scale link. Secondly, the boundary layer is more mixed, as discussed earlier, and the mean potential temperature gradient is even slightly negative closest to the surface. It is interesting to study the energy fluxes across the surface in these two distinctly different conditions, as illustrated in Figure 7. The cloudy and clear episodes are clearly indicated by the probability function for the net longwave radiation at the bottom of the plot, as two separate peaks. The top panel shows the contributions to the surface energy balance by the two other remaining heat fluxes, the turbulent heat flux (essentially the sensible heat flux as the flux water-vapour flux is very small) and the conduction of heat through the surface (the differences in sign is due to the definition as positive upwards).

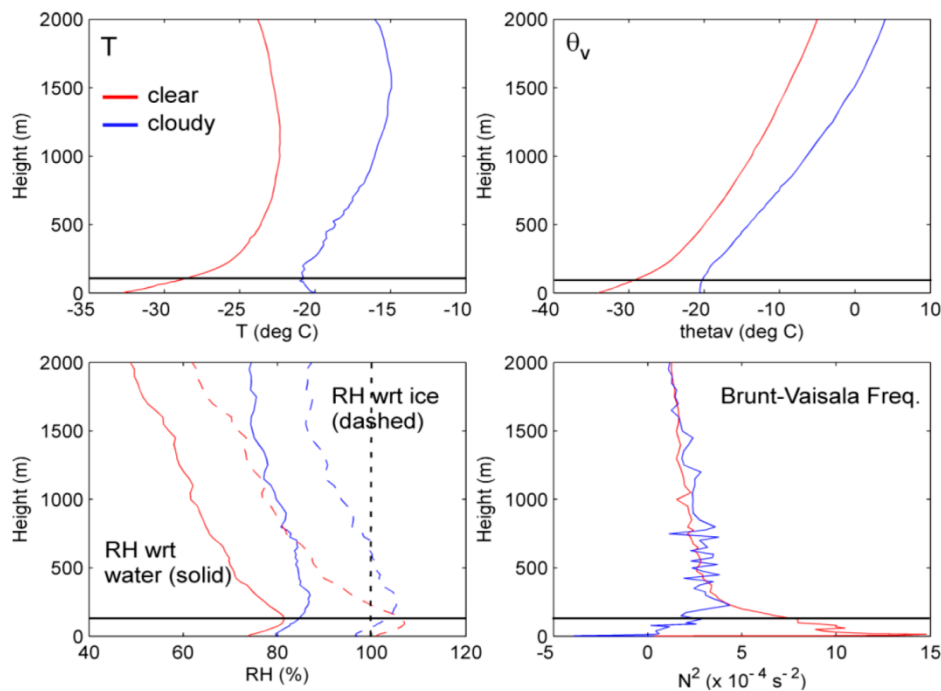


Figure 6. Winter time mean vertical profiles of temperature and virtual temperature (top panels, °C), relative humidity (lower left, %) and stability, expressed as the Brunt Vaisalla frequency squared (lower right, s^{-2}) for soundings from SHEBA divided into clear (red) and cloudy (blue) cases. Reproduced by courtesy of Ola Persson.

In clear conditions the surface loses $\sim 40 \text{ Wm}^{-2}$ of energy due longwave radiative cooling and this partly compensated in approximately equal measures by both the turbulent heat flux and by conduction of heat through the ice and snow. Both are directed towards the surface and each is slightly more than $\sim 10 \text{ Wm}^{-2}$ on average; the net average energy balance is -17.5 Wm^{-2} and the surface is thus cooling.

When it is cloudy, on the other hand, the net radiation is close to zero. However, a large temperature difference between the ocean water at the underside of the ice and the atmosphere on top of course remains and therefore the conduction of heat through the ice continues to be upward and thus heat the surface. Consequently the turbulent heat flux must change sign and becomes a small but mostly upward flux compensating by cooling the surface; the net energy flux at the surface is very small, about $\sim 1.5 \text{ W m}^{-2}$ on average, and this state is thus a quasi-steady state. Figure 7 thus explains quite well both the episodic behaviour of the surface temperature in Figure 4a and the differences in vertical structure associated with the clouds and clear conditions in Figure 6.

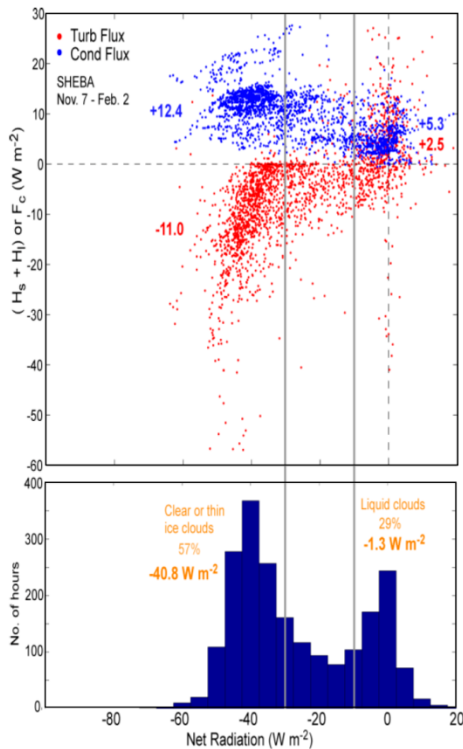


Figure 7. The surface energy balance in winter showing the PDF of net long-wave radiation in the bottom panel and the turbulent and conductive heat fluxes in the top panel. See the text for a discussion. Reproduced by courtesy of Ola Persson.

5. The summer-time Arctic boundary-layer

Figure 8 and 9 show the typical summer time (July and August) lower troposphere structure from four expeditions (AOE-96, SHEBA, AOE-2001 and ASCOS), illustrated by the statistics of the equivalent potential temperature and the relative humidity (wrt liquid). Quite clearly the conditions are mostly near neutral in the lowest few hundreds of meters. The corresponding statistics for relative humidity shows that this layer is on average very moist, always $RH > 95\%$ and often $\sim 100\%$. A peculiarity is that the moist layer is often deeper than the near-neutral layer. This indicates that the specific humidity must increase across the capping inversion; if the relative humidity is constantly near saturation and the temperature increases, then the specific humidity must also increase. This condition is prevalent in the Arctic and occurs also in winter (e.g. Sedlar et al. 2011b) and as a consequence entrainment across the capping inversion is a source of moisture, opposite to what is typical across stratocumulus in other regions. The persistently very high relative humidity in the boundary layer is likely due to the combination of a surface near saturation and an entrainment of humidity across the capping inversion. While Sedlar et al. (2011b) examines this in detail they fail to explain why the

moisture increases with height across the inversion, but speculates it is related to the advection of moist and warm air from lower latitudes. Another consequence of the high relative humidity is of course a persistent occurrence of low-level clouds. While the total cloud fraction is high in summer, typically around 90%, Figure 10 shows the statistics of the lowest cloud-base height and of the thickness of the lowest cloud layer. Quite obviously, very low stratocumulus or stratus clouds dominate. The turbulent heat fluxes are typically small (not shown), well within $\pm 10 \text{ Wm}^{-2}$, and the surface energy flux is dominated by radiation which in turn is largely controlled by clouds.

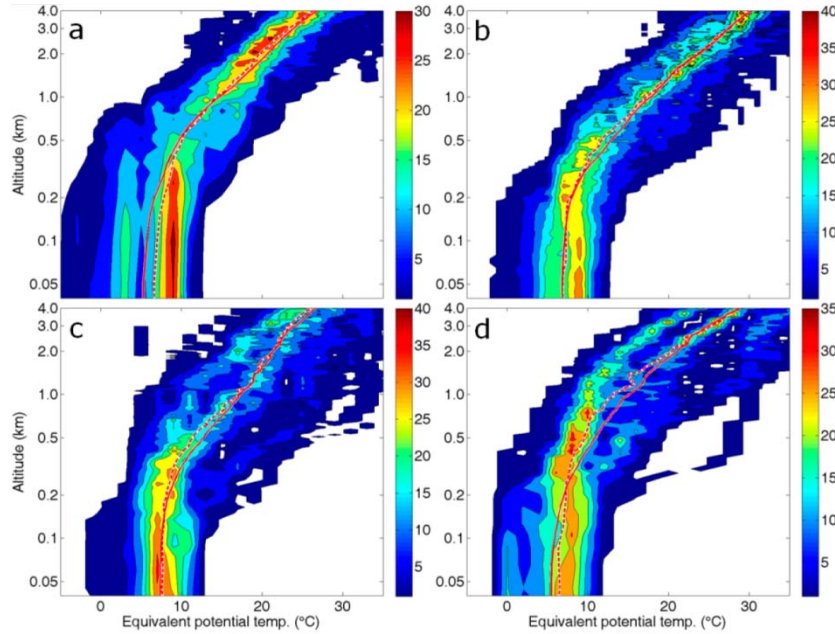


Figure 8. The probability of the equivalent potential temperature during four different summers: (a) ASCOS in 2008, (b) AOE-2001, (c) SHEBA in 1998 and (d) AOE-96.

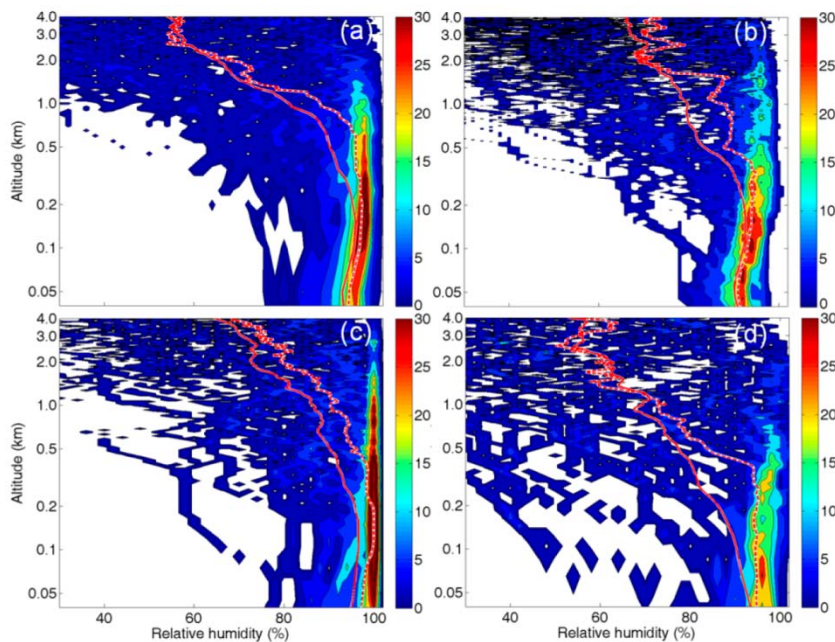


Figure 9. Same as Figure 8 but for relative humidity

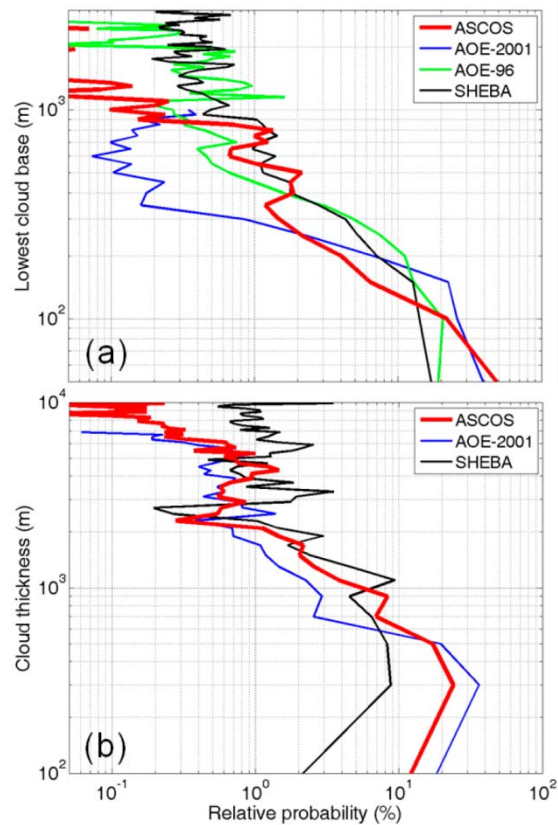


Figure 10. Statistics of the lowest cloud-base height and the thickness of the lowest cloud layer from the expeditions in Figure 8-9.

Another peculiarity in the Arctic summer PBL is that aerosol concentrations are typically quite low; median total aerosol concentration during ASCOS was around 100 cm^{-3} and 10 cm^{-3} were common. One manifestation of this is the apparent absence of haze as indicated by visibility observations (not shown); even with very low cloud bases the visibility outside fog is very seldom below 20 km. A consequence of this is that the number concentration of cloud condensation nuclei (CCN), regulating the number of cloud droplets, is often low, sometimes so low that they effectively impedes cloud formation (Mauritsen et al. 2011).

This results in optically thin clouds, almost invisible to the eye but readily detected by for example laser cloud ceilometers. An example from the ASCOS expedition is shown Figure 11. The cloud layer becomes thin around the midnight between the two days but is still detected as a low cloud between the surface and 150 m by the combination of radar and laser ceilometer. Visual observations, on the other hand, show blue sky aloft. Looking away from the sun, however, is a very distinct bow-like optical phenomenon indicative of the presence of very large spherical droplets. Also at very low angles to the horizon the cloud layer can be seen, but looking upward or towards the sun the clouds appear absent to the eye. As a consequence of these conditions, increasing the aerosol concentration induces a relative warming of the surface since the increased longwave radiation far outweighs the decreased solar radiation at the surface (e.g. Mauritsen et al. 2011). Analysis of CCN observations from four late summer Arctic expeditions on the Swedish icebreaker Oden (IAOE-91, AOE-96, AOE-2001 & ASCOS) indicates that such conditions may appear as frequent as ~20-30% of the time.

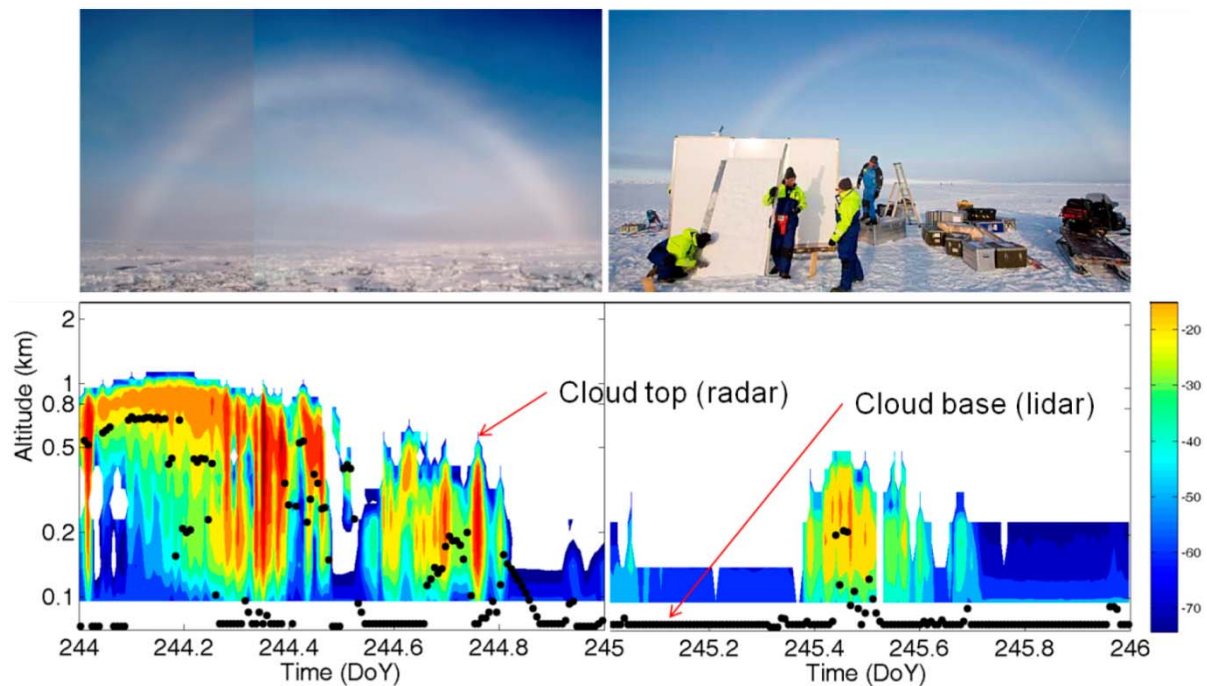


Figure 11. A two-day episode from ASCOS, showing the evolution of a cloud layer using cloud radar reflectivity (color), with the lowest cloud base height from a laser ceilometer indicated by black solid dots. When the radar reflectivity indicates presence of hydrometeors below the cloud base, this is due to precipitation. On top are two photographs taken away from the sun during the thin cloud episode around midnight (photographs by Michael Tjernström, left, and Thorsten Mauritsen, right).

6. Some modelling problems

It appears evident from the previous sections that the Arctic Ocean atmospheric conditions are a challenge to numerical weather prediction or climate models. Indeed, several studies have indicated significant Arctic modelling problems (Walsh et al. 2002; Tjernström et al. 2005, 2008; Rinke et al. 2006; Wyser et al. 2008; Karlsson and Svensson 2010). In Tjernström et al. (2005) the ARCMIP (Curry and Lynch 2002) simulations of several regional climate models were compared to observations from SHEBA; later Tjernström et al. (2008) used the same model results to examine the clouds. Here we will only address two specific problems: the effect of clouds on surface radiation and the turbulent heat fluxes. Tjernström et al. (2005) concluded that while using a relatively small regional model domain and forcing with European Centre for Medium Range Weather Forecasts (ECMWF) operation analyses at the boundaries, the large scale meteorology was well constrained and that for basic meteorological variables such as mean sea-level pressure 2-m temperature and 10-m winds all models performed acceptably, although wind speed was somewhat worse than temperature and pressure.

Tjernström et al. (2008) calculated the effects of clouds on incoming radiation at the surface, longwave and shortwave separately. Incoming radiation was examined used instead of net radiation to isolate the effects of the clouds and to avoid complications from errors in how the surface is treated. While modelled surface radiation has a relatively high correlation to observed radiation it is clear that there is a negative bias in both long- and shortwave radiation reaching the surface (not shown).

Figure 12 illustrates the results, showing the zenith-angle corrected shortwave radiation and the apparent sky emissivity as a function of the vertically integrated cloud-water content, in the models and from SHEBA. Note that this representation does not say anything about how well the cloud water is calculated in the models, which is in general quite poor (not shown). For the most typical cloud water paths, solar radiation is clearly under-predicted by almost all models while the incoming longwave radiation is also underestimated in several models.

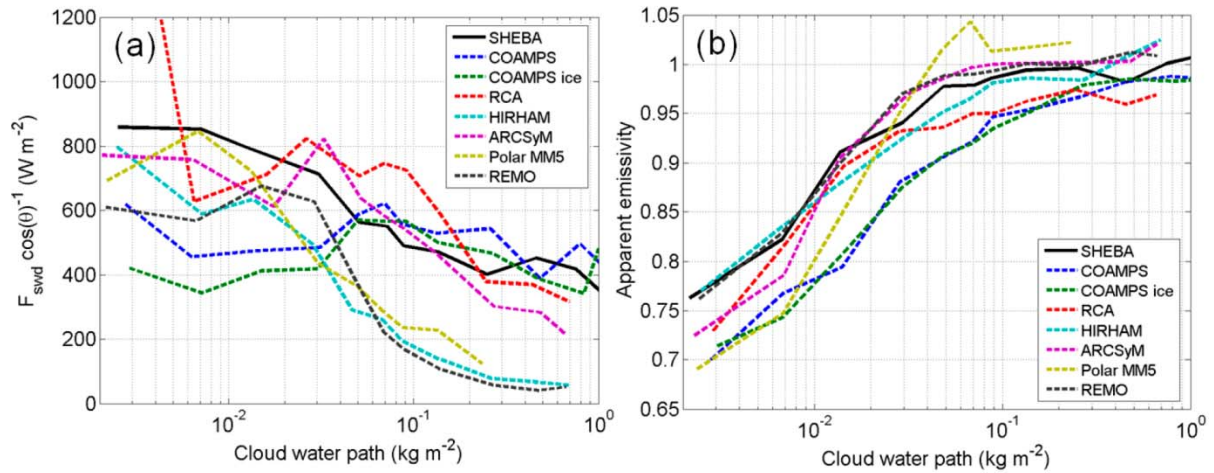


Figure 12. Plots of the effects of clouds on the incoming radiation showing (a) incoming solar radiation and (b) the apparent emissivity from several regional climate models for the ACMIP simulations and for SHEBA, as functions of cloud water (liquid plus ice) path. Incoming solar radiation is normalized for varying zenith-angle and the apparent emissivity is calculates as the ratio of the incoming longwave radiation to the blackbody radiation with respect to the near-surface air temperature. Reproduced from Tjernström et al. (2008).

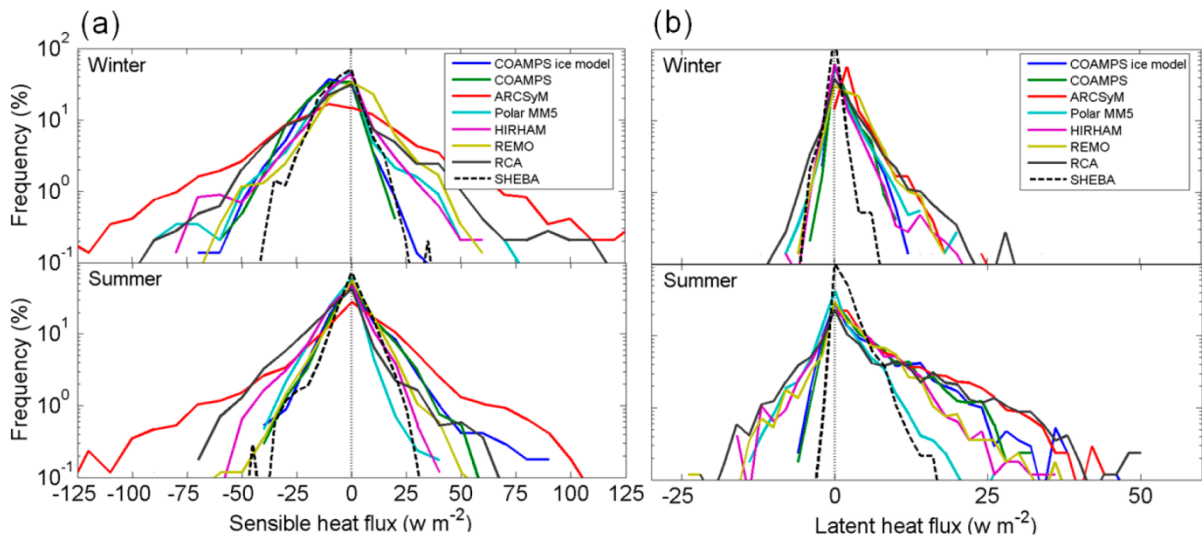


Figure 13 .The probability functions for the turbulent (a) sensible and (b) latent heat fluxes (Wm^{-2}) for the ARCMIP regional climate model simulation, stratified into the winter and summer symmetric half year.

Tjernström et al. (2008) speculated on two different causes relating to the dominant effects of the clouds in different seasons. Solar radiation is attenuated too much by the clouds; the clouds are to optically thick especially for the stratiform low-level clouds, with low or moderate cloud-water paths. Since the analysis is stratified by cloud-water path, systematically to thick or thin clouds in the models are cancelled out. Another potential reason could be the cloud microphysics assumptions in the models, not accounting for the low CCN concentrations that are often observed thereby making the clouds to opaque to solar radiation. For longwave radiation also winter conditions are included; none of the models have any substantial liquid water in the cold winter conditions. Stratifying the winter conditions into clear and cloudy conditions, Tjernström et al. (2008) could show that while the error for clear conditions was scattered around zero in all models, there was a substantial negative bias for cloud conditions that could not be explained by errors in cloud temperature (e.g. cloud base height). Thus the conclusion is that the failure to account for the presence of liquid water in mixed-phase clouds was the main cause for the underestimation of incoming longwave radiation.

Tjernström et al. (2005) also examined the turbulent heat fluxes at the surface for the ARCMIP simulation. The momentum flux (not shown) was overestimated in all models on average, such that there are too many occasions with a large momentum flux and too few with a small. It was also found that the magnitudes of the turbulent heat fluxes were always overestimated, regardless of sign, by anywhere from a factor of two to four. It is likely that the overestimation of the turbulent heat fluxes comes as a consequence of the overestimation of the momentum flux, since these are intimately coupled in the models. Accumulated over a season these errors lead to huge overestimations in the heat fluxes, however, for the atmospheric modelling it turns out that very large errors in accumulated downward sensible heat flux is compensated by similarly huge errors in accumulated upward latent heat flux (evaporation); the conclusion is that such model results cannot be tolerated if coupling the atmospheric to ice and ocean models. Sorteberg et al. (2007) examined the heat fluxes in a large number of the coupled global models in the CMIP3 archive, and in ECMWF ERA-40 and NCEP reanalyses, and also found very much too large turbulent fluxes, especially for the latent heat flux.

7. Conclusions

The remote Arctic Ocean with its perennial sea-ice surface that is melting in summer, with its large annual cycle holds a very special climate regime that challenges modelling in many different ways. A full understanding of these properties relies on very few sets detailed observations mostly from shorter expeditions and a climatology of the boundary-layer characteristics from the Arctic Ocean does not exist due to lack of observations. Accurate consideration of boundary-layer processes and the vertical structure of the lowest troposphere without proper consideration of the both the special surface properties of the ice covered ocean and the impacts on the surface radiation by the clouds.

Some properties of this special climate regime stand out:

- In an annual context, the boundary-layer structure is dominated by the near-neutral conditions, but strong long-lasting surface inversions occur in autumn and in winter. Near-neutral well-mixed boundary layers totally dominates in spring and summer, but occur also in autumn and winter somewhat less than half the time. The boundary layer is predominantly shallow, most often less than one or two hundred meters.

- The boundary layer changes abruptly between the two states: strong surface-based inversions and well-mixed near-neutral stratification. The changes in structure between these two states are dictated by the presence of low clouds.
- Cloudiness is high, near 90% in summer and 60-70% in winter and low-level clouds dominates. Due to the fact that the surface albedo is comparable to the cloud albedo, these clouds are warming the surface, compared to cloud-free conditions, by their “greenhouse effect” except possibly during summer when melting of snow and ice opens up low-albedo areas of open water and forms darker melt ponds of fresh water on the surface of the ice.
- The low clouds are often mixed-phase with a layer of liquid water near the cloud top, even in very cold conditions, quasi-continuously precipitating ice particles. While low clouds dominate, these clouds are sometimes optically thin. In summer this is because of sometimes very low aerosol concentrations, impeding formation of CCN, while liquid-water is important in winter presumably due to a paucity of freezing nuclei.
- Regardless of boundary-layer stratification, absolute humidity often increases with height across the boundary-layer inversion, likely due to long-range transport of moist air from lower latitude. This contributes to the high boundary-layer relative humidity by entrainment of moisture down from the lower free troposphere.
- Three surface coupling regimes exist: 1) The summer melt season, with a nearly-fixed non-responsive surface temperature with respect to surface net radiation, around the melting points of saline ocean water (~ -1.8 °C) and fresh water (snow on surface; ~ 0 °C); 2) Non-melt but solar forced spring and autumn seasons, with sub-zero temperatures responding to radiative forcing and a clear diurnal cycle; 3) Polar night with sub-zero responsive temperatures, fluctuating between clear and cloudy due to longwave forcing.
- Near-surface moisture remains high, almost always close to saturation with respect to liquid water in summer but with respect to ice in the colder season. This reduces relative humidity with respect to liquid water in the cold season and is partly responsible for the lower cloud fraction in the cold season.
- The surface energy balance is dominated by the radiation fluxes. The longwave radiation is the only factor in the polar night but also often dominates in summer, due to the high albedo of the snow and ice.
- While the other surface energy fluxes (turbulent fluxes and conduction of heat from the ocean under the ice) are smaller than the net radiation in magnitude, they are still important in shaping the boundary layer. In winter, conduction through ice and snow is as important as the sensible heat flux and the thickness of the snow on the surface of the ice is critical to this flux. The latent heat flux is negligible in winter except from open leads.

To properly model these conditions requires a high vertical resolution near the surface, to resolve strong vertical temperature gradients in the winter and the shallow boundary layer in all seasons, as well as the low clouds stratiform clouds. Also required for the energy balance is adequate cloud modeling, for the radiation fluxes, a realistic specification of the ice thickness and modeling of the highly variable snow thickness, to properly resolve the conduction of heat through the ice and snow, and an accurate representation of the stable boundary layer. Adequate cloud modeling probably requires some specification of aerosol concentrations to properly account for the frequent occurrences of low concentrations of cloud condensation nuclei in summer and of ice nuclei in all seasons.

Acknowledgements

This work was sponsored by the Swedish Natural Research Council, the Knut and Alice Wallenberg Foundation, the DAMOCLES European Union fp6 Integrated Project, the Swedish Polar Research Secretariat and the Bert Bolin Centre of Climate Research. The author is indebted to the many scientists and logistics staff responsible for, or contributing to, the collection of data at the SHEBA ice camp and from the four Swedish Arctic atmospheric expeditions on the icebreaker *Oden*. Special thanks to Ola Persson, Matthew Shupe, Thorsten Mauritsen, Ian Brooks, Cathryn Birch and Joseph Sedlar.

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